ILMATIETEELLISEN KESKUSLAITOKSEN TOIMITUKSIA

FINNISH METEOROLOGICAL OFFICE CONTRIBUTIONS

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SOME COMPUTATIONS OF THE ENERGY EXCHANGE BETWEEN THE SEA AND THE ATMOSPHERE IN THE BALTIC AREA

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Helsinki 1964. Valtioneuvoston kirjapaino

1. Introduction

Although information concerning the energy exchange between the sea and the atmosphere is of great significance both theoretically and practically, the present methods for computing the exchange are still relatively inaccurate. For the Baltic Sea computations or estimations of the mean energy exchange have been made by several authors, e.g. WITTING (1918), JOHNSON (1940), SIMOJOKI (1946 and 1949), HELA (1951) and BROGMUS (1952). JOHNSON has made an attempt to determine the whole heat balance of the Baltic Sea, but in the main these studies only take into account evaporation. Recently PALMÉN (1963) has calculated evaporation from the Baltic Sea for the period of a certain year, using the divergence of the atmospheric vapour flux.

The first main purpose of this study is to compare PALMÉN's evaporation equation with the empirical formula (1) of JACOBS (1942), and the second is to estimate the most significant terms in the annual heat budget of the Bothnian Sea (the southern part of the Gulf of Bothnia), using observations of the Swedish lightship »Finngrundet» (61°04′N, 18°41′E). Computations of evaporation and of the flux of sensible heat over the Baltic Sea proper were made for the period December 1961—May 1962. For the Bothnian Sea, furthermore, the long- and short-wave radiation and the heat used to melt solid precipitation were estimated for the period of one year March 1961—February 1962.

2. Methods used

For computing the evaporation the familiar formula of the type (JACOBS, 1942):

$$E = k\,(e_s - e_a)\,V_a \tag{1}$$

was used. Here e_s denotes the saturated water vapour pressure at the temperature of the water surface, e_a the water vapour pressure of the air at the altitude a and V_a the wind velocity at the same altitude.

The values of k in Eq. (1) used by several authors in general range from 0.09 to 0.16, when E is expressed in mm $(24h)^{-1}$, e in mb and V_a in

ms^{-1}, and when the height a is about 6 m. At least the following values of k have been used: MASUZAWA (1952) 0.093, SVERDRUP (1945) 0.099, SCHULEIKIN (1932) (according to STURM, 1962) 1.00, SAMOILENKO (1952) (according to STURM, 1962) 0.105, SIMOJOKI (1949) 0.110, PRIVETT (1960) 0.114, JACOBS (1951) 0.143 and MIKHALEVSKII (according to LAEVASTU, 1960) 0.157. MARCIANO and HARBECK (1954) tested several evaporation equations and found that SVERDRUP's (1937) theoretical equation gave results that were satisfactory but somewhat too large. SVERDRUP's formula can be expressed in the form of Eq. (1), where the coefficient k varies between 0.108 and 0.120 in the conditions existing in the Baltic area. According to PETTERSSEN *et al.* (1962), SWINBANK (1959), combining observations made by SHEPPARD (1958) and MARCIANO and HARBECK (1954), found that a value of about 0.130 gave reasonable results. Most of the above-mentioned constants were originally intended to be applicable only to climatic data, but many of them have also been used for short periods by several authors. PRIVETT's value of 0.114, which is very close to the average of the foregoing values, was chosen for the present calculations.

The height a in Eq. (1) was assumed to be 6 m, but during regular observations $\bar{e_a}$ was measured at 2 m and V_a at about 10 m. On »Finngrundet» both were measured at 4 m altitude. The correction necessitated by this fact was estimated, using the measurements of WÜST (1937) concerning the vertical distribution of water vapour pressure above the sea, and the well known logarithmic wind law

$$(2) \qquad \frac{V_a}{V_b} = \frac{ln \frac{a}{z_o}}{ln \frac{b}{z_o}}$$

where a and b are the heights above the sea surface and z_o the »roughness parameter». The quantity $\frac{e_s - e_a}{e_s - e_b}$ must also be a constant, if e is a linear function of $ln\ z$. According to WÜST (1937) and MONTGOMERY (1940), this is the case for the most part, especially with unstable stratification of the air. Using 26 cases of WÜST's measurements, where $e_s - e_2 > 0$ in each case, and the values of MARCIANO and HARBECK (1954) for z_o, it was found that on the average

$$\frac{\bar{e}_s - \bar{e}_6}{\bar{e}_s - \bar{e}_2} \frac{\overline{V}_6}{\overline{V}_{10}} = 1.02; \quad \frac{\bar{e}_s - \bar{e}_6}{\bar{e}_s - \bar{e}_4} \frac{\overline{V}_6}{\overline{V}_4} = 1.11$$

Thus, no significant error will be caused by using the regular observation heights instead of 6 m. For the Baltic Sea calculations the evaporation equation

(3a) $$E = 0.114\ (e_s - e_2)\ V_{10}\ \mathrm{mm(24h)^{-1}}$$

was used. For computing the evaporation on »Finngrundet» the coefficient in Eq. (3a) was multiplied by 1.11:

(3b) $$E = 0.127\ (e_s - e_4)\ V_4\ \mathrm{mm(24h)^{-1}}.$$

The effect of salinity on e_s has been left out of account. According to SIMOJOKI (1946), the mean salinity of the surface waters of the Northern Baltic Sea is 5.5—6.5 per mille, and according to SVERDRUP (1945) this would cause a decrease of less than 4 per mille in the water vapour pressure.

Some authors, e.g. MONTGOMERY (1940) and SVERDRUP (1946), have assumed that the sea surface is changed from smooth to rough by a wind speed of approximately 6.5 ms^{-1}, whereat the rate of evaporation changes discontinuosly. Thus the value of k in Eq. (1) with a smooth surface would be different from that with a rough one. This is doubted by several authors. MARCIANO and HARBECK (1954), for example, concluded from careful observations on Lake Hefner that the free water surface is always to be considered as rough. The same value of k has been used for all wind speeds in this paper.

Using the mean value 593 cal g^{-1} for the latent heat of vaporization ($L_T = 597 - 0.544\ T_s$ cal g^{-1}), we obtain the heat used for evaporation (the flux of latent heat) corresponding equations (3a) and (3b) as follows:

(4a) $$Q_e = L_T E = 6.75\ (e_s - e_2)\ V_{10}\quad \mathrm{cal\ cm^{-2}\ (24\ h)^{-1}}$$

(4b) $$Q_e = L_T E = 7.50\ (e_s - e_4)\ V_4\quad \mathrm{cal\ cm^{-2}\ (24\ h)^{-1}}$$

According to BOWEN (1926), the relation between the convective transfer of sensible heat and the flux of latent heat is

(5) $$R = \frac{Q_c}{Q_e} = \frac{c_p\, p}{L_T\ 0.623}\ \frac{T_s - T_a}{e_s - e_a} \backsim 0.65\ \frac{p}{1\,000}\ \frac{T_s - T_a}{e_s - e_a},$$

where c_p is the specific heat of air at constant pressure, p the atmospheric pressure (mb), T_s the temperature of the sea surface (°C) and T_a the temperature of the air at the altitude a (°C). Combining equations (3a) and (5) and using the value 1 000 mb for p, we obtain the formula:

(6a) $$Q_c = 4.40\ (T_s - T_2)\ V_{10} \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1}$$

which has been used for computing Q_c on the Baltic Sea. Similarly, combining equations (3b) and (5), we obtain the formula:

(6b) $$Q_c = 4.88\ (T_s - T_4)\ V_4 \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1}$$

for computing the flux of sensible heat on »Finngrundet». Authors who have used formulae of type (6) have used at least the following coefficients (T_a and V_a measured at 6 m): PRIVETT (1960) 2.29, SAMOILENKO (1959) (according to STURM, 1962) 3.45, KANGOS (1960) 5.53 and PETTERSSEN *et al.* (1962) 6.77.

For estimating the short- and long-wave radiation the formulae recommended by LAEVASTU (1960) were used. LAEVASTU's formula for the diurnal total (solar and diffuse) short-wave radiation is

(7) $$Q_s = 0.014\ A_n t_d\ (1 - 0.0006\ C^3) \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1},$$

where A_n is the noon altitude of the sun (degrees), t_d the length of the day from sunrise to sundown (minutes) and C the cloudiness (scale 1—10).

The radiation reflected from the sea surface is, according to LAEVASTU (1960),

(8) $$Q_r = 0.15\ Q_s - (0.01\ Q_s)^2 \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1}.$$

For computing the effective back radiation (long-wave radiation) in a clear sky, LÖNNQUIST's (1954) formula in the following reduced form given by LAEVASTU (1960) was used:

(9) $$Q_{ob} = 297.0 - 1.86\ T_s - 0.95\ U_o \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1}.$$

U_o is here the relative humidity (per cent). Q_{ob} is corrected for the effect of cloudiness with MÖLLER's (1953) formula given by LAEVASTU (1960)

(10) $$Q_b = Q_{ob}\ (1 - 0.0765\ C) \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1}.$$

Q_b is consequently the real effective back radiation.

An attempt was also made to compute the effect of the melting of solid precipitation (snow or hail) by using the formula

(11) $$Q_p = 8.0\ P \quad \text{cal cm}^{-2}\ (24\ \text{h})^{-1},$$

were P is the solid precipitation in mm of water per day. The temperature of the snow is not taken into account.

The net energy flux from the sea surface was determined from the formula

$$Q_n = Q_e + Q_c + Q_b + Q_p + Q_r - Q_s. \tag{12}$$

Figure 1 shows the area of computation for the Baltic Sea proper. There are 21 grid points in the area marked with letters from A to V.

The sea surface temperature was measured on board s/s »Ariadne», which runs the route Helsinki—Copenhagen once a week (see fig. 1). Observations on board are made every fourth hour, and during each run 6—8 observations were made within the area of computation. Temperatures for the intermediate days were interpolated. On the northern border of the area (points A, B and C), however, sea surface temperatures observed on Bengtskär (59°43′N, 22°30′E), were used. At those grid points at which ice sometimes occurred during the winter, the value $T_s = -0.3$ °C (the approximately freezing temperature) was used when ice occurred. From the the southeastern Baltic there were no actual observations of sea surface temperatures. During the period December—March no systematic temperature differences between the different parts of the Baltic Sea were obtained, and the mean temperature of each run of s/s »Ariadne» was used for the whole area, except the points mentioned above. In April the southern Baltic began to warm up faster. From April 1 the area was divided into two parts, with the latitude 57°30′N as the border and temperatures measured from s/s »Ariadne» were averaged separately for the two parts.

The evaporation and the flux of sensible heat were computed for the Baltic Sea proper twice a day, i.e. using synoptic observations made at 00 GMT and at 12 GMT. Observations on the temperature and the water vapour pressure of the air, as well as on the wind speed from nearly all the meteorological stations in the Baltic area (see fig. 1), were plotted on the maps. The fields of T_2, e_2 and V_{10} were carefully analysed. There are no ship observations available from the Baltic Sea, therefore the weight in in the analyses were put on the observations made at the few really maritime stations to avoid the continental influence upon the observations. Thus, for example, in temperature analyses, especially with light winds, a relatively equal temperature field was assumed to exist on the open sea, while the isotherms were very close near the coast. At those grid points where fast ice occurred, the values $E = Q_c = 0$ were used.

The mean evaporation in the whole area was computed from the formula

(13)

$$\bar{E} = \frac{A + C + D + G + N + S + V + 2\,(B + H + J + K + M + R + T + U)}{44}$$

$$\frac{+\,3\,(E + F + O) + 4\,(I + L + P)}{44},$$

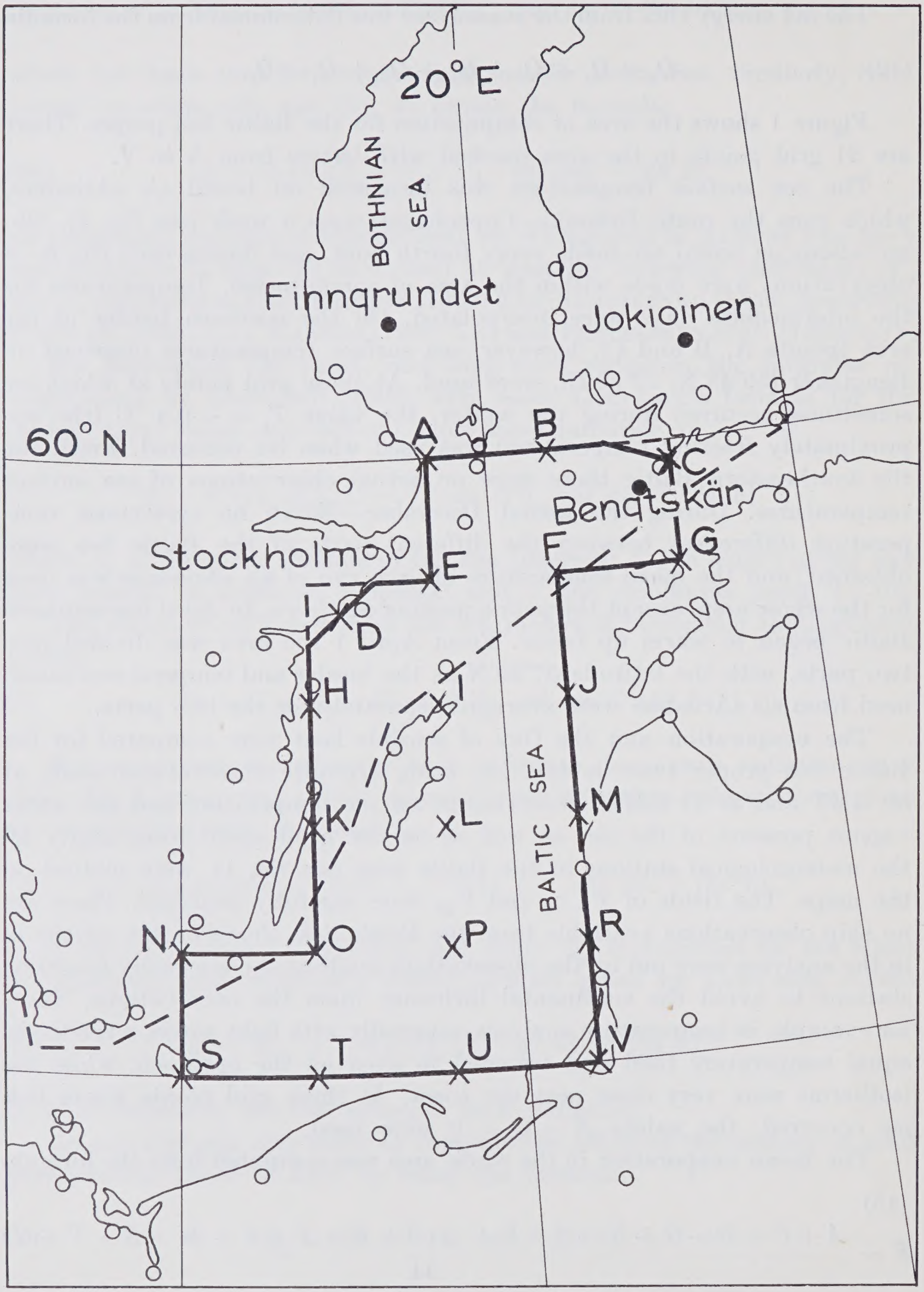

Fig. 1. The area of computation (heavy line), the grid points used and the locations of some places mentioned in the text. The observation stations used for the Baltic Sea computations are marked with small open circles. The dashed line indicates the approximate route of s/s »Ariadne».

where the letters from A to V now indicate the evaporation at the corresponding points. The mean value of Q_c was evaluated in the same manner.

Winter 1960—1961 was mild in Scandinavia, and the Bothnian Sea, which normally freezes in late winter, did not freeze at all. The next winter the sea froze during the first week of March. It was therefore reasonable to choose the period March 1961—February 1962 for the computations, because the heat storage of the Bothnian Sea was at an approximate minimum at both ends of the period.

On the lightship »Finngrundet» the sea surface temperature is observed twice a day (at 06 and 18 GMT) and the meteorological observations are made three times a day (at 06, 12 and 18 GMT). Thus no observation is made at night. For computing the diurnal values of E, Q_c, Q_{ob} and Q_b (equations 3b, 6b, 9 and 10) all the meteorological observations were smoothed according to the formula:

$$\bar{O} = \frac{2(O_6 + O_{18}) + O_{12}}{5}, \tag{14}$$

where $\bar{O}$ is the daily mean of any observation. O_6, O_{12} and O_{18} are the observations made at the corresponding GMT.

The effect of the ship itself upon the observed values of the temperature and water vapour pressure of the air has been neglected. This subject has been discussed in greater detail by HELA (1951), who came to the conclusion that on »Finngrundet» the effect of the ship is negligible.

Table 1. The monthly amounts of evaporation E, fluxes of latent and sensible heat Q_e and Q_c, the BOWEN ratio R, the total energy exchange $Q_a = Q_e + Q_c$ and evaporation according to PALMÉN on the Baltic Sea during the period December 1961—May 1962.

Month	E mm	Qe cal cm-²	Qc cal cm-²	R	Qa cal cm-²	PALMÉN E mm
D	**103.5**	**6 130**	**4 675**	**0.76**	**10 805**	**94**
J	49.7	2 962	2 013	0.68	4 975	67
F	62.5	3 683	2 523	0.69	6 206	42
M	39.0	2 314	1 533	0.66	3 847	27
A	12.8	758	—1 181	— 1.56	— 423	19
M	*8.9*	*530*	*—1 192*	— *2.3*	— *662*	*8*
Sum	276.4	16 377	8 371		24 748	257

3. Results

The results from the Baltic Sea proper are presented in table 1. Both the evaporation (103.5 mm) and the flux of sensible heat (4 675 cal cm^{-2}) were plainly largest in December. In December the total energy exchange

$Q_a = Q_e + Q_c$ was 10 805 cal cm^{-2}. After a mild autumn the Baltic Sea had an unusually high surface temperature in December, while the temperature of the air was colder than normal. Hence these amounts of energy flux are larger than the average in December. In April and May the evaporation was relatively small, as would be expected. Both the flux of sensible heat and the total energy flux were directed from the atmosphere to the sea during these months. During the whole period the total energy flux was about 24 700 cal cm^{-2}.

The proportion of sensible heat is relatively great in winter. The monthly BOWEN ratio R (computed from the monthly values of Q_c and Q_e) ranges from 0.66 to 0.76 in the months December—March. According to JACOBS (1951), the mean value of R in the Northern Atlantic between the corresponding latitudes 55° and 60°N in winter is about 0.45.

It is understandable that in winter the BOWEN ratio is greater over the Baltic Sea than over the ocean at the same latitudes, although the relative humidity is probable greater over the ocean. The temperature of the sea surface in winter is far lower in the Baltic Sea, and at low temperatures the saturation value of e decreases very slowly as the temperature falls. Therefore the difference $e_s - e_a$ (Eq. 5) cannot reach such very high values as the difference $T_s - T_a$ does in winter, especially during aoubreaks of cold air. Momentarily R exceeded the value 1 comparatively often. The highest momentary values about 1.3, when both Q_c and Q_e were positive.

Besides the foregoing the different methods used by JACOBS and the author require to be considered. JACOBS used the mean values of any element, i.e. the BOWEN ratio is proportional to

$$\frac{\overline{T}_s - \overline{T}_a}{\bar{e}_s - \bar{e}_a} = \frac{(\overline{T}_s - \overline{T}_a)\,\overline{V}_a}{(\bar{e}_s - \bar{e}_a)\,\overline{V}_a}.$$

In this paper R is computed from monthly values of Q_c and Q_e, i.e. R is proportional to

$$\frac{\overline{(T_s - T_a)\,V_a}}{\overline{(e_s - e_a)\,V_a}} = \frac{(\overline{T}_s - \overline{T}_a)\,\overline{V}_a + \overline{(T_s - T_a)'\,V_a'}}{(\bar{e}_s - \bar{e}_a)\,\overline{V}_a + \overline{(e_s - e_a)'\,V_a'}}.$$

Now it is obvious that in winter the latter expression gives greater values. For the major part of the energy flux (in form of Q_c and Q_e) occurs during outbreaks of cold air, when R reaches its highest values, i.e. the quantities $\overline{(T_s - T_a)'\,V_a'}$ and $\overline{(e_s - e_a)'\,V'}$ play a noteworthy role. JACOBS has used climatological mean values and the author monthly means, it is true, but in principle this does not change the situation.

When using the BOWEN ratio (Eq. 5), it is assumed that the eddy exchange coefficients for water vapour and heat are identical. According to the turbulence theory, the vertical flux of heat is

$$Q_c = - c_p A_{cz} \frac{\partial \Theta}{\partial z} \backsim - c_p A_{cz} \frac{\partial T}{\partial z} \tag{15}$$

and of water vapour

$$Q_e = - A_{ez} \frac{\partial q}{\partial z} = - \frac{0.623}{p} A_{ez} \frac{\partial e}{\partial z} \cdot \tag{16}$$

Here A_{cz} is the vertical eddy exchange coefficient (the eddy conductivity for heat, A_{ez} the vertical eddy exchange coefficient (the eddy diffusivity) for vapour, Θ the potential temperature, q the specific humidity of the air and z the vertical coordinate. However, it is not evident that A_{cz} and A_{ez} are equal. For instance, according to PASQUILL (1949), over open grasslands A_{cz} is reasonably equal to A_{ez} under stable conditions, but is substantially greater in unstable conditions and may even be twice as great. If this is also true over a water surface, the amounts of Q_c would be relatively greater, maybe even greater than those of Q_e over the Baltic Sea in winter, when the air is mostly unstable.

In table 1 the evaporation values for the same period according to PALMÉN (1963) are also presented. (In this table the value 3 mm for May has been corrected to 8 mm according to a verbal suggestion by Prof. PALMÉN). The agreement can be regarded as fair. In general the values of PALMÉN are somewhat smaller but in January and May plainly larger. PALMÉN's value of 257 mm for the period of half a year is about 7 per cent smaller than the corresponding value of 276 mm obtained by the author. In comparing the results, the method of PALMÉN and the possible errors in both calculations will be discussed.

PALMÉN has used aerological data and the observed precipitation for computing the mean evaporation from the formula

$$\bar{E} - \bar{P} = \frac{1}{g} \int_{p_h}^{p_o} \frac{\partial \bar{q}}{\partial t} dp + \frac{L}{gA} \int_{p_h}^{p_o} \widehat{qV_n} \, dp. \tag{17}$$

Here A is the horizontal area, L the length of its periphery, $\bar{P}$ the mean precipitation over the area, g the gravity constant, p_o the surface pressure of the air, p_h the pressure at the upper boundary of the volume (here 400 mb), $\bar{q}$ the mean specific humidity at a given isobaric surface and $\widehat{qV_n}$ the mean

horizontal flux of water vapour normal to the boundary L. The area A, however, was somewhat larger than that used in this paper.

Assuming the flux of liquid and solid water carried by clouds to be of minor importance, PALMÉN's method can be considered relatively exact. It is probable, however, that in special cases of strong evaporation, especially during the cold season, the neglect of liquid and solid water in the clouds will result in an underestimate of the amount of evaporation. MANABE (1958) has made an attempt to estimate this error when the cold air in winter moves from the Asian continent over the Japan Sea. According to MANABE, the magnitude of the error of the computation of the horizontal divergence of water vapour is about 14 % in the cloudy layer between the coast of the Asian continent and the northwest coast of the Japanese islands. Over the Japan Sea, however, the conditions are far more extreme and more liable to error of this kind than over the Baltic Sea. Thus it may be roughly estimated that PALMÉN's evaporation values for the winter months are a few per cent too small. Further, small observation errors and the long distances between the aerological stations used may cause considerable errors in the evaporation values, but these errors will probably counterbalance each other during a sufficiently long period. PALMÉN's value for February is obviously too small, because for some days during a strong outbreak of cold air observation data were lacking, and PALMÉN has used the mean daily evaporation values of February for these days. The magnitude of this error may be about 10 mm.

On the assumption that evaporation can be expressed by a formula of type (1), the major error sources for computing the evaporation in this study will be as follows:

(a) Eq. (3a) represents evaporation in mean conditions. It is probable, however, that the vertical flux of water vapour (and the coefficient k in Eq. 1) is also dependent on the stability of the air. MANABE (1958) has estimated the dependence of k on the air stability from the energy budget, and KONDO (1962) has made a theoretical attempt to evaluate evaporation and the vertical flux of sensible heat as a function of stability. Both authors conclude that k in Eq. (1) will be greater in unstable than in stable conditions. The same applies to the coefficient in equations of type (6). The present author, however, has intended to estimate only the mean value of k.

(b) Because no ship observations were available, the values of e_a (and T_a) over the sea were possibly too low in winter and too high in spring, in spite of the careful analyses mentioned in section 2. As a result, the values of E (and Q_c) would be slightly too large in winter and slightly too small in spring. In addition, the value for the whole period would be somewhat too large. It is impossible to make any exact estimate of the magnitude of this error, but it is probably relatively insignificant.

Table 2. The monthly amounts of evaporation E, fluxes of latent and sensible heat Q_e and Q_c, the Bowen ratio R, the total energy exchange $Q_a = Q_e + Q_c$, the effective back radiation Q_b, heat used to melt solid precipitation Q_p, the total incoming radiation Q_s, reflection back from the sea surface Q_r and the net flux of energy from the sea surface Q_n on »Finngrundet» during the period March 1961—February 1962.

Month	E mm	Qe cal cm-2	Qc cal cm-2	R	Qa cal cm-2	Qb cal cm-2	Qp cal cm-2	Qs cal cm-2	Qr cal cm-2	Qn cal cm-2
M	32.7	1 940	— 885	— 0.46	1 055	3 560	30	6 670	865	— 1 160
A	13.6	810	— 765	— 0.94	45	**4 155**	35	12 420	1 345	— 6 840
M	*1.0*	*60*	*—1 100*	*—18.3*	*—1 040*	3 205	—	16 220	1 580	—12 475
J	3.3	195	—1 080	— 5.5	— 885	3 215	—	**19 845**	1 655	*—15 860*
J	17.8	1 050	— 285	— 0.27	765	2 990	—	18 180	**1 665**	—12 760
A	49.4	2 905	— 300	— 0.10	2 605	2 940	—	13 115	1 410	— 6 160
S	55.1	3 240	— 265	— 0.08	2 975	3 270	—	7 815	895	— 675
O	53.0	3 120	— 145	— 0.05	2 975	2 805	—	3 480	435	2 735
N	69.8	4 130	1 340	0.32	5 470	3 045	—	1 325	195	7 385
D	**87.4**	**5 180**	**3 710**	0.72	**8 890**	3 455	70	*605*	*90*	**11 900**
J	57.3	3 405	2 555	**0.75**	5 960	2 995	130	860	130	8 355
F	54.5	3 235	2 330	0.72	5 565	*2 550*	**230**	2 230	310	6 425
Year	494.9	29 270	5 110	0.17	34 380	38 185	495	102 765	10 575	—19 130

(c) The s/s »Ariadne» observations of T_s were made on the open sea, while some of the grid points (fig. 1) are located over coastal waters. In consequence, the values of E (and Q_c) will also be too large in winter and too small in late spring. The magnitude of this error was estimated as follows. The mean monthly temperatures of the sea surface given by Böhnecke and Dietrich (1951) were considered. Those grid points where the mean monthly temperatures on the s/s »Ariadne» route (within the computation area) were nearly valid, were regarded as representative ones, as well as points A, B and C (the Bengtskär values) and the points where ice occurred. The other points were regarded as non-representative. The mean temperature differences (according to Böhnecke and Dietrich, 1951) between the representative and non-representative grid points were assumed to remain the same during the months considered here. According to these calculations, the evaporation values of each month must be corrected by the following values:

December —7.3 mm	February —1.9 mm	April 0.0 mm
January —2.7 mm	March —0.4 mm	May +0.3 mm

As a result, the evaporation of the whole period would be reduced by 12.0 mm.

On combining the foregoing estimates, it will be found that the evaporation value of the whole period computed from Eq. (3a) is about 264 mm or a little smaller. Palmén's value for the same period is nearly the same, if

Table 3. The monthly air temperatures (T_a) at Mariehamn (60°07′N, 19°54′E) and the monthly temperatures of the surface water (T_s) on »Finngrundet» during the period March 1961—February 1962 in comparison with the normal values.[1])

Month	Ta °C			Ts °C		
	actual	normal	diff.	actual	normal	diff.
M	2.0	— 1.7	+3.7	0.9	*0.4*	+0.5
A	3.3	2.7	+0.6	1.5	1.3	+0.2
M	8.4	7.9	+0.5	3.5	3.0	+0.5
J	**15.2**	12.5	+2.7	8.9	7.2	+1.7
J	15.0	**16.5**	—1.5	13.1	13.6	—0.5
A	13.8	15.6	—1.8	**13.7**	**15.2**	*—1.5*
S	11.4	11.3	+0.1	12.5	11.8	+0.7
O	10.6	6.3	**+4.3**	11.1	7.9	**+3.2**
N	3.9	2.7	+1.2	5.8	5.2	+0.6
D	—2.6	—0.2	*—2.4*	3.1	3.2	—0.1
J	—1.3	—2.8	+1.5	1.7	1.4	+0.3
F	*—2.7*	*—4.1*	+1.4	*0.6*	0.7	—0.1
Year	6.4	5.5	+0.9	6.4	5.9	+0.5

the February value is corrected, as was suggested. If noticeable net flux of liquid and solid water from the area exist, PALMÉN's value must be somewhat increased. In comparing the results of PALMÉN (1963) and the author, it can be concluded that PRIVETT's (1960) coefficient 0.114 in Eq. (3a) on the average gives good or possibly somewhat too small evaporation values.

The results from »Finngrundet» are presented in table 2. All monthly evaporation values were positive. The smallest monthly evaporation, 1.0 mm, occurred in May and the largest, 87.4 mm, in December. The total annual evaporation was about 495 mm. The flux of sensible heat also reached its lowest monthly value, —1 100 cal cm^{-2}, in May and its highest, 3 710 cal cm^{-2}, in December. According to the monthly values it was directed from the sea to the atmosphere in only four months, from November to February, but the annual value, 5 110 cal cm^{-2}, was plainly positive. Table 3 shows that in May and October the air temperature was abnormally high in the Baltic area. Therefore it can be assumed that normally Q_c is positive in these months. The total energy exchange by convective processes, Q_a, was positive in every month except May and June. During the period December—February, when computations were made for both the Baltic Sea and »Finngrundet», both E and Q_c were considerably smaller on »Finngrundet» in December and February but larger in January. In the period October—February evaporation on »Finngrundet» was 322 mm and over the Baltic Sea proper, according to PALMÉN (1963), 321 mm.

[1]) The normal values of T_s are according to BÖHNECKE and DIETRICH (1951). The normal T_s value for March is only approximative.

Table 4. Mean monthly evaporation according to BROGMUS and HELA, mean monthly flux of sensible heat according to HELA and the amounts computed by the present author (table 2) on the Bothnian Sea.

Month	BROGMUS E mm	HELA E mm	HANKIMO E mm	HELA Qc cal cm-²	HANKIMO Qc cal cm-²
J	46	37	57	1 443	2 555
F	39	43	55	1 410	2 330
M	22	36	33	916	— 885
A	16	13	14	— 835	— 765
M	4	*—15*	*1*	—1 508	*—1 100*
J	*1*	—11	3	*—1 648*	—1 080
J	12	13	18	—1 162	— 285
A	28	**61**	49	— 203	— 300
S	46	57	55	297	— 265
O	42	53	53	1 027	— 145
N	57	37	70	1 340	1 340
D	**62**	37	**87**	**1 808**	**3 710**
Year	375	361	495	2 885	5 110

Values of E and Q_c from table 2 are presented in table 4 with the mean values for the Bothnian Sea computed by HELA (1951) and BROGMUS (1952). HELA has used the formulae of DEVIK (1932)

$$(18) \qquad Q_c = 0.0439 \; T_K \sqrt{V_a + 0.3} \; (T_s - T_a) \quad \text{cal cm}^{-2} \; (24 \text{ h})^{-1},$$

$$(19) \qquad Q_e = 72.5 \; \frac{T_K}{p} \sqrt{V_a + 0.3} \; (e_s - e_a) \quad \text{cal cm}^{-2} \; (24 \text{ h})^{-1},$$

where T_K is the absolute temperature of the sea surface. BROGMUS (1952) has used Eq. (1), where the coefficient k varies from 0.110 to 0.118, when the same units are used as in Eq. (3a). Both HELA and BROGMUS have obtained considerably smaller values than the present author.

Although the values of BROGMUS are smaller, the annual rate of evaporation is nearly the same as that given here: the largest values occur in December and November and the smallest ones in May and June. According to HELA, on the contrary, the strongest evaporation occurs in August and September. This is probably due to the fact that in Eq. (19) evaporation is proportional to the square root of V_a but in Eq. (1) to V_a, for the mean wind velocity is far greater in winter than in summer. WITTING (1918) has estimated the mean annual evaporation over the Bothnian Sea at 425 mm and SIMOJOKI (1949) at 421 mm. Thus the annual evaporation of 495 mm obtained here is considerably larger than the means estimated by the abovementioned authors.

In the months December—February the BOWEN ratio ranged from 0.72 to 0.75 and was hence of exactly the same magnitude as on the Baltic Sea. Here, too, the highest daily values of R were about 1.3, when both Q_c and Q_e were positive. The annual value 0.17 shows that the contribution of Q_c to the annual energy budget is rather small. HELA (1951) obtained a mean annual value of 0.134 for R on »Finngrundet». JACOBS (1951) found the annual mean over the North Atlantic at latitude 60°N to be about 0.24. Apart from the different methods used to compute R, the difference is probably due to the fact that over a limited sea as Baltic the amounts of Q_c in summer overcompensate for the greater values of R in winter.

The monthly values of the effective back radiation from the sea surface shown in table 2 range from about 2 500 to 4 200 cal cm^{-2}. The largest values occurred in early spring, when the cloud cover and the relatively humidity were smallest. Daily values of Q_b varied between 45 and 215 cal cm^{-2}. The annual value, about 38 200 cal cm^{-2}, is somewhat larger than the annual Q_a. According to SIMOJOKI (1946) the mean annual value of Q_b at Bogskär (59°31′N, 20°31′E) computed with DEVIK's (1932) method is about 41 200 cal cm^{-2}.

As shown in table 2, the monthly and annual amounts of heat used to melt snow are unimportant in the energy budget. Only on one single day Q_p was the greatest term in the energy budget.

According to table 2, the sea should have received a net heat increase of about 19 000 cal cm^{-2} year^{-1}, but according to observations of the sea temperature the sea has lost energy during the year in question. An obvious reason for this disagreement is LAEVASTU's (1960) method (equations 7 and 8) for computing the incoming short-wave radiation, which gives too large values. As table 5 shows, the actual Q_s values measured in Jokioinen and Stockholm (see fig. 1) are considerably smaller. Particularly the term 0.014 $A_n t_d$ in Eq. (7), which express the short-wave radiation from a clear sky, seems to affect the error. According to LUNELUND (1940), the mean annual total incoming radiation at the latitude of »Finngrundet» is about 109 000 cal cm^{-2} and that computed from Eq. (7) about 135 000 cal cm^{-2}, which is about 24 % larger than the former value. It was therefore necessary to estimate Q_s and Q_r in another way.

The radiation values measured in Jokioinen (table 5) were chosen as the basic ones. It can be assumed that on the average the same air mass is lying over the Bothnian Sea and Jokioinen. Therefore the mean turbidity of the air (except the cloud cover) will be nearly the same over the two places. »Finngrundet» had a smaller mean cloud cover in every month except September and is located 15′ further north than Jokioinen. These facts were taken into account by correcting the monthly radiation values of Jokioinen as follows:

Table 5. The monthly values of the total incoming radiation measured at Stockholm and at Jokioinen and values computed from Eq. (7) during the period March 1961— February 1962.

Month	Stockholm Qs cal cm-2	Jokioinen Qs cal cm-2	Computed Qs cal cm-2
March	6 699	5 348	6 670
April	11 600	10 817	12 420
May	11 330	11 543	16 220
June	**14 776**	**14 516**	**19 845**
July	11 191	11 247	18 180
August	10 124	9 125	13 115
September	6 436	6 053	7 815
October	3 024	2 033	3 480
November	1 034	416	1 325
December	*693*	*270*	*605*
January	953	422	860
February	2 113	1 492	2 230
Year	79 973	73 282	102 765

$$Q_{sF} = K \frac{(1 - 0.0006\ C_F^3)\ Q_{sJ}}{1 - 0.0006\ C_J^3} \tag{20}$$

Here K is the relationship between the mean monthly short-wave radiation with a clear sky at the latitude of »Finngrundet» and that at the latitude of Jokioinen, according to LUNELUND (1949). The subscripts F and J signify »Finngrundet» and Jokioinen; other notations are as in Eq. (7).

For estimating Q_r the following reflexion values (in per cent) were used: from April to August 7 %, in March and September 10 %, in February, October and November 14 %, in December and January 20 %. These values are somewhat subjectively chosen on the basis of ANDERSON's (1954) studies concerning reflection from water at various sun altitudes and cloud covers.

Table 6 presents the corrected monthly values for Q_s and Q_r as well as the net heat fluxes from the sea surface. The annual Q_s value, about 78 300 cal cm^{-2}, is larger than the Q_s measured in Jokioinen and a little smaller than that measured in Stockholm and may be regarded as fairly representative. The corresponding value 102 300 cal cm^{-2} computed from Eq. (7) is about 31 % larger. According to table 6, the sea would have lost about 1 000 cal cm^{-2} during the year in question. Thus the annual heat budget of the Bothnian Sea seems to be excellently balanced. However, this may be considered a mere accident.

The mean depth of the Bothnian Sea is about 70 m (according to WITTING, 1908). According to measurements made on »Finngrundet» and on some other vessels, the heat storage of the sea in a column 70 m × 1 cm² has

Table 6. The corrected monthly values of the total incoming radiation Q_s, reflection back from the sea surface Q_r and the net flux of energy from the sea surface Q_n on »Finngrundet» during the period March 1961—February 1962.

Month	Qs cal cm-²	Qr cal cm-²	Qn cal cm-²
March	6 000	600	— 755
April	11 310	790	— 6 285
May	12 350	865	— 9 320
June	**14 680**	**1 030**	*—11 320*
July	12 560	880	— 7 925
August	10 560	740	— 4 275
September	5 950	595	890
October	2 120	300	3 960
November	490	70	8 095
December	*270*	*55*	**12 200**
January	520	105	8 670
February	1 480	210	7 075
Year	78 290	6 240	1 010

decreased by approximately 4 500—5 000 cal during the year. The difference between the observed and computed heat loss, 3 500—4 000 cal cm^{-2}, will be compensated for, in the main, by advection of heat from the vicinity of »Finngrundet» and/or by inaccuracies in the computed energy budget terms. Other terms in the energy budget are so small that they can be ignored. If the heat transported by fresh-water discharge is included in the advection term, the largest of these small terms is heat from the dissipation of wind and tidal energy. According to Laevastu (1960), in the Irish channel this amounts to about 1 000 cal cm^{-2} year^{-1}. In the Baltic area the corresponding influence of tides must be very much smaller.

Figure 2 shows the annual course of the energy storage per cm² in the Bothnian Sea according to the monthly Q_n values from table 6. The heat storage on March 1. has been regarded as zero. According to fig. 2, the heat storage in 1961 reached its minimum in the middle of March and the maximum occurred in early September. The annual fluctuation from the first minimum (March 1961) to the maximum was about 40 500 cal cm^{-2} and from the maximum to the second minimum (March 1962) approximately 41 500 cal cm^{-2}. Johnson (1940) has estimated the corresponding value in the Baltic Sea at 44 500 cal cm^{-2}.

It may be asked whether the values computed for »Finngrundet» are valid over the whole Bothnian Sea. »Finngrundet» is located in the southern part of the Bothnian Sea, and at least the annual values of incoming short-wave radiation and evaporation in general decrease northwards. It is there-

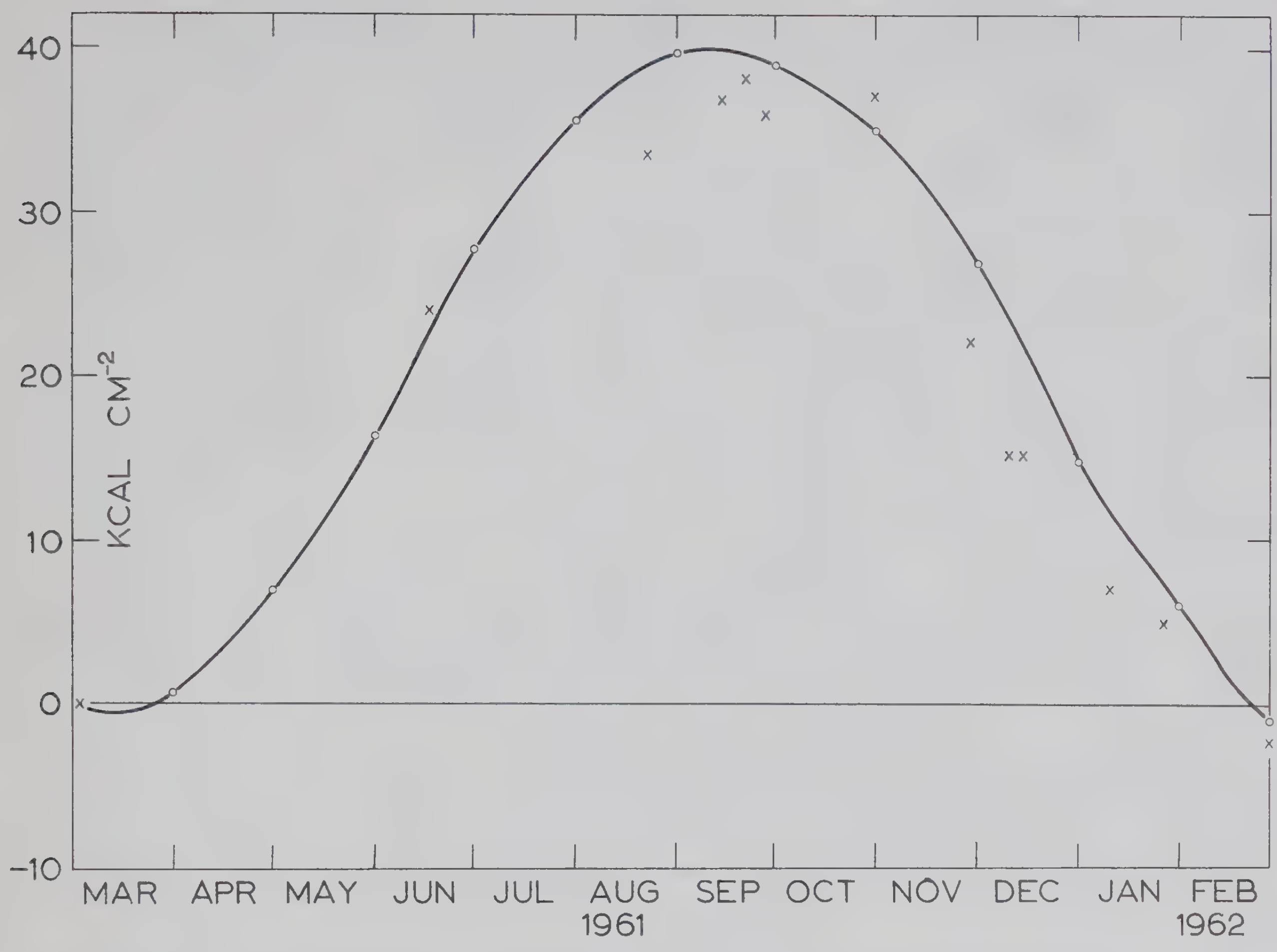

Fig. 2. The annual course of the heat storage per cm² in the Bothnian Sea, according to the »Finngrundet» computations (solid line) and according to the measurements made at the point 61° 05′ N, 19° 35′ E (the crosses). The heat storage at the beginning of the period has been taken as zero.

fore obvious that the annual variation of the heat storage in the whole Bothnian Sea is somewhat smaller. Furthermore, the size of the Bothnian Sea is a noteworthy factor: a small sea has a relatively larger area of coastal waters, where the conditions for energy exchange are quite different from those over the open sea.

The crosses in figure 2 show the heat storage in a column 40 m × 1 cm² at the point 61° 05′ N, 19° 35′ E. The value for June 17 is observed at the point 61° 50′ N, 20° 06′ E, and the values at both ends of the period are approximated from the measurements made on »Finngrundet», where the sea temperature was observed to the depth of 30 m. Here, likewise, the heat storage at the beginning of the period has been regarded as zero. The specific heat of water was taken as 1 cal $g^{-1}\,{}^{\circ}C^{-1}$ and the density of water as 1 g cm^{-3}. The thermocline in the Bothnian Sea mostly occurs above the 40 m level; thus the above-mentioned water column represents the annual course of the heat storage relatively well, although the mean depth in the Bothnian Sea is about 70 m. The two heat storage estimates show relatively good agreement.

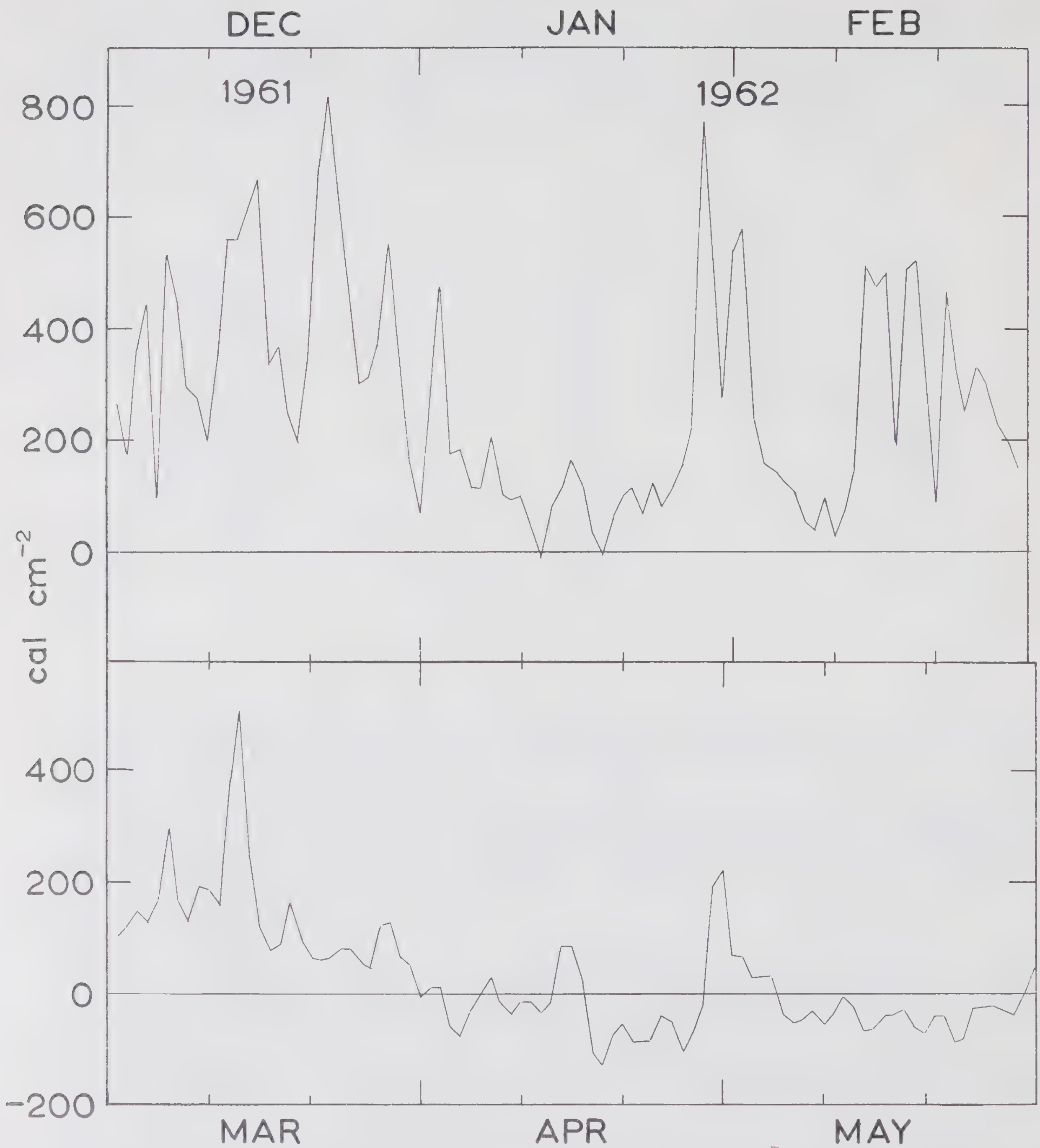

Fig. 3. The diurnal amounts of the total energy lost from the sea surface by convective processes ($Q_a = Q_e + Q_c$) on the Baltic Sea proper.

The crosses are considerably dispersed, however, which is at least partly due to the advection of heat. For example, there has been strong positive advection between the end of August and the end of October. These advections are due rather to the water moving in the Bothnian Sea itself than to exchange between the surrounding seas. In November the heat storage decreased according to the crosses much more than according to the solid line (fig. 2). This is at least partly due to vertical mixing. For according to

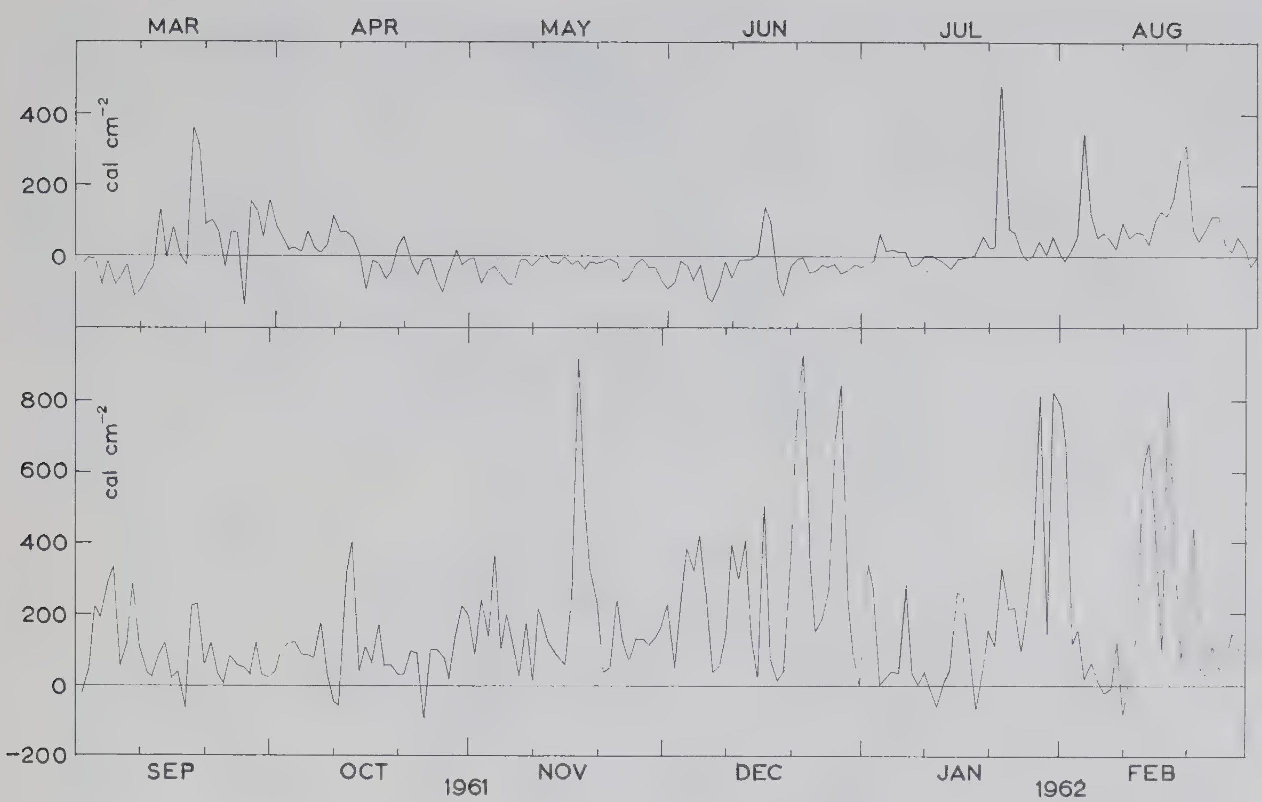

Fig. 4. The diurnal amounts of the total energy lost from the sea surface by convective processes ($Q_a = Q_e + Q_c$) on »Finngrundet».

Simojoki (1946), it is in November that a very strong heat flux occurs owing to conduction from the layers above 40 m into the deeper layers in the Baltic Sea.

Figures 3 and 4 show the diurnal amounts of Q_a over the Baltic Sea and on »Finngrundet». It will be seen that in late spring and early summer the flux of heat was comparatively steady from day to day and mostly directed from the atmosphere to the sea. In winter, on the other hand, there were large variations in the daily values. The highest peaks indicate strong outbreaks of cold air, and almost the whole monthly energy flux might be concentrated in a few days, as for example in February on »Finngrundet». The curve for the Baltic Sea is somewhat smoothed in comparison to that for »Finngrundet», owing to the relatively large area.

In winter the amounts of incoming radiation were quite insignificant (table 6), and Q_b also varied relatively little in comparison to Q_a (the difference between the largest and smallest diurnal value of Q_b was 170 cal cm^{-2}). Hence the Q_a curve provides a comparatively good representation of the relative variations of the net energy flux from the sea in winter, i.e. during the period November—February. In late spring and early summer, on the contrary, the variations in the diurnal Q_s values are naturally the greatest in the energy budget.

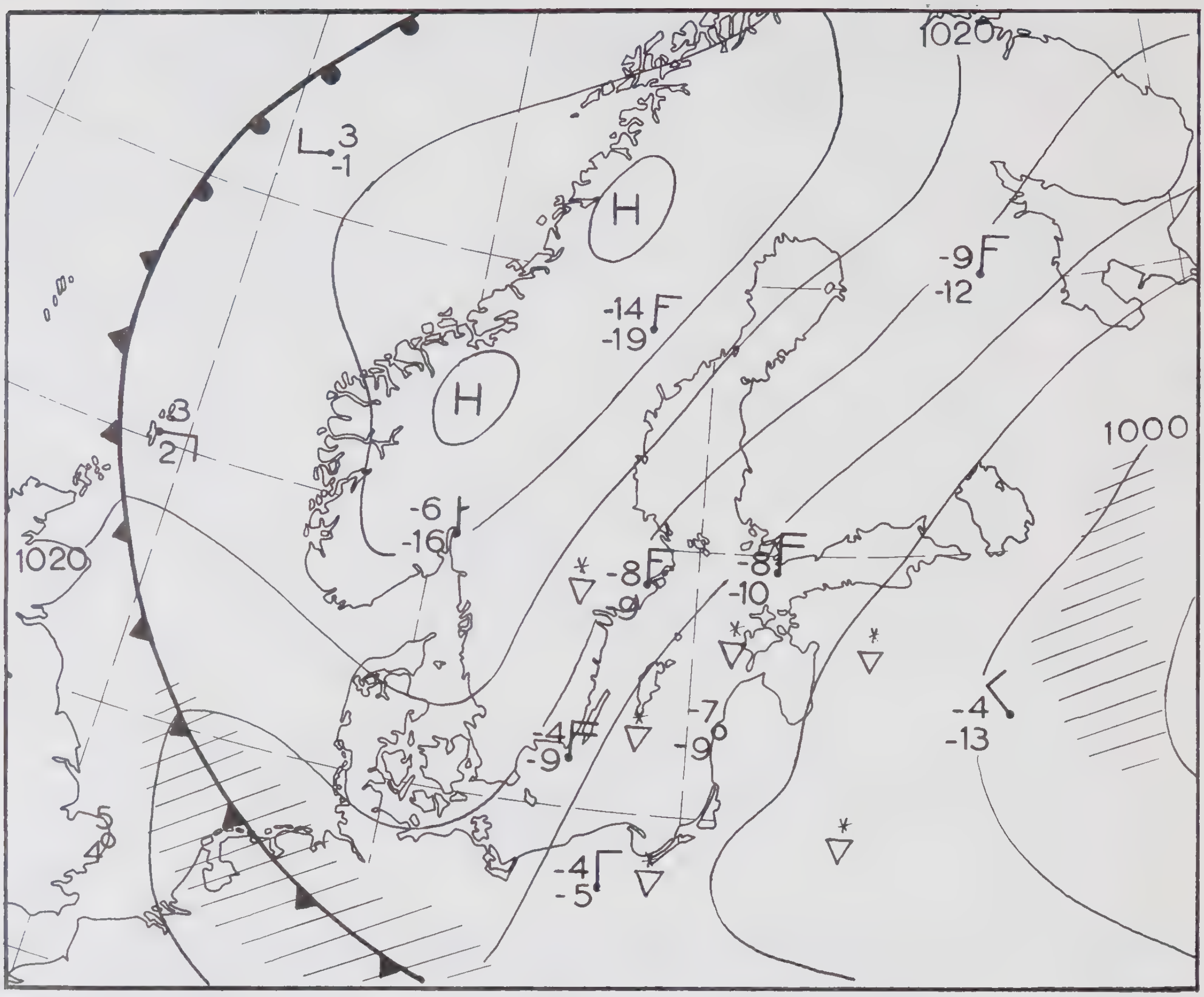

Fig. 5. Synoptic surface map on Dec. 22, 12 GMT 1961. A typical situation of strong energy flux from the sea to the atmosphere. Diurnal values on the Baltic Sea: E 6.0 mm, Q_a 732 cal cm^{-2}; on »Finngrundet»: E 9.0 mm, Q_a 935 cal cm^{-2}.

The largest diurnal values of Q_a occurred both on the Baltic Sea (732 cal cm^{-2}) and on »Finngrundet» (935 cal cm^{-2}) during an outbreak of cold air from the north on the December 22 (fig. 5). The latter value is equal to 0.65 cal cm^{-2} min^{-1} or about one-third of the solar constant. Over the oceans the corresponding energy fluxes may be far larger, e.g. according to Petterssen *et al.* (1962) in extreme cases Q_a may exceed 2.5 cal cm^{-2} min^{-1} over small regions of the North Atlantic. The smallest diurnal value of Q_a was —114 cal cm^{-2} on the Baltic Sea and —145 cal cm^{-2} on »Finngrundet». On the Baltic Sea the values of Q_c varied between —96 and 375 cal cm^{-2} $(24\ h)^{-1}$ and on »Finngrundet» between —115 and 475 cal cm^{-2} $(24\ h)^{-1}$. The largest diurnal evaporation was 6.3 mm on the Baltic Sea and 11.0 mm on »Finngrundet». Such strong diurnal evaporation as Palmén (1963) has found, was not obtained on the Baltic Sea, and even on »Finngrundet» the

daily evaporation exceeded 7 mm only twice during the year. The fact that PALMÉN has obtained far larger (somewhat over 10 mm) daily evaporation values during some outbreaks of cold air seems to support the assumption that coefficient k in Eq. (1) is greater in unstable conditions. The smallest diurnal evaporation was —0.3 mm on the Baltic Sea and —0.6 mm on »Finngrundet». Thus the amounts of condensed water vapour computed from equations (3 a) and (3 b) are quite small. Also it is not obvious that the same evaporation equation is valid for both evaporation and condensation, but the error in the annual heat budget due to this fact is quite insignificant, because the real amounts of condensation are evidently also very small.

4. Conclusion

1. In comparing the results obtained by PALMÉN (1963) and by the author it was concluded that the evaporation values given by PRIVETT's coefficient $k = 0.114$ in Eq. (1) were on the average satisfactory but perhaps somewhat too small. The water vapour pressure was measured at 2 m and the wind velocity at 10 m, but the same coefficient would obviously also have been valid, if both e_a and V_a had been measured at 6 m.

2. The annual evaporation on the Bothnian Sea was estimated at 495 mm during the year in question. The mean annual value may be somewhat smaller, because at the end of the year the heat storage of the sea was smaller than at the beginning.

3. The proportion of sensible heat was quite small in the energy exchange except in winter, when it was on the average 70—75 per cent of the flux of latent heat.

4. LAEVASTU's (1960) formula (7) for the total incoming radiation was found in this case to give values about 30 per cent too large in comparison with the actual measurement.

5. When the computed values of incoming radiation were replaced by the measured ones, the heat budget of the Bothnian Sea (Eq. 12) was found to be satisfactorily balanced.

6. The total heat loss from the maximum to the minimum annual heat storage of the sea according to Eq. (12) was estimated at about 41 000 cal cm^{-2} on »Finngrundet». For the Bothnian Sea as a whole the corresponding value is probably somewhat smaller.

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References

Anderson, E. R., 1954: Energy-budget studies. Water loss-investigations: Lake Hefner Studies. U. S. Geol. Survey, Prof. Pap., 269, pp 71—119.

Böhnecke, G. and G. Dietrich, 1951: Monatskarten der Oberflächentemperatur für die Nord- und Ostsee und die angrenzenden Gewässer. Deutsches Hydr. Inst. Hamburg.

Bowen, I. S., 1926: The ratio of heat losses by conduction and by evaporation from any water surface. Phys. Rev., Ser. 2, vol. 27, pp 779—787.

Brogmus, W., 1952: Eine Revision des Wassershaushaltes der Ostsee. Kieler Meeresforschungen, Bd. IX, H. 1, pp 15—42.

Devik, O., 1932: Thermische und dynamische Bedingungen der Eisbildung in Wasserläufen. Geofys. Publ., vol IX.

Hela, I., 1951: On the energy exchange between the sea and the atmosphere in the Baltic area. Ann. Acad. Sci., Fennica, Ser. A. 1 (97), pp 1—48.

Jacobs, W. C., 1942: On the energy exchange between sea and atmosphere. J. Mar. Res., vol. V, No 1, pp 45—47.

—»— 1951: The energy exchange between sea and atmosphere and some of its consequences. Bull. Scripps Inst. Oceanogr., 6 (2), pp 27—122.

Johnson, N., 1940: Östersjöns värmeekonomi. Sv. Hydr. Biol. Komm. Skr., Ny serie: Hydr. XV.

Kangos, J. D., 1960: A preliminary investigation of heat flux from the oceans to the atmosphere in Antarctic regions. J. Geophys. Res. 65 (12), pp 4007—4012.

Kondo, J., 1962: Evaporation from extensive surfaces of water. Sci. Rep. Tohoku Univ., Ser. V, vol. 14 (3), pp 107—119.

Laevastu, T., 1960: Factors affecting the temperature of the surface layer of the sea. Soc. Sci. Fennica, Comm. Phys.-Math., XXV (1), pp 1—136.

Lönnquist, O., 1954: Synthetic formulae for estimating effective radiation to a cloudless sky and their usefulness in comparing various estimation procedures. Ark. Geofys., 2 (12), pp 247—294.

Lunelund, H., 1940: In Finnland eingestrahlte Wärmemengen. Soc. Sci. Fennica, Comm. Phys.-Math. XI (2), pp 1—19.

Manabe, S., 1958: On the estimation of energy exchange between the Japan Sea and the atmosphere during winter based upon the energy budget of both the atmosphere and the sea. J. Met. Soc. Japan, Ser. II, vol. 36, No 4, pp 123—134.

Marciano, J. J. and G. E. Harbeck jr, 1954: Mass transfer studies. Water loss-investigations: Lake Hefner Studies. U.S. Geol. Survey, Prof. Pap., 269, pp 46—70.

Masuzava, J., 1952: On the heat exchange between sea and atmosphere in the southern Sea of Japan. Ocean. Mag. 4 (2), pp 49—55.

Möller, F., 1953: Tabellen zur atmosphärischen Strahlung und Optik. Linke's Meteorologisches Taschenbuch, Bd. 2, Leipzig.

MONTGOMERY, R. B., 1940: Observations of vertical humidity distribution above the ocean surface and their relation to evaporation. Pap. phys. Oceanogr. and Meteor., vol. 7, No 4.

PALMÉN, E., 1963: Computation of the evaporation over the Baltic Sea from the flux of water vapor in the atmosphere. I. A. S. H. Comm. Evap., No 62, pp 244—252.

PASQUILL, F., 1949: Eddy diffusion of water vapour and heat near the ground. Roy. Soc. London, Ser. A, vol. 198, No 1052 (22), pp 116—140.

PETTERSSEN, S., D. L. BRADBURY and K. PEDERSEN, 1962: The Norwegian cyclone models in relation to heat and cold sources. Geofys. Publ. Geophysica Norwegica, vol. XXIV, pp 243—280.

PRIVETT, D. W., 1960: The exchange of energy between the atmosphere and the oceans of the southern hemisphere. Geophys. Mem. 104, pp 3—61.

SAMOILENKO, V. S., 1952: The available theory of oceanic evaporation and its practical application (russ.). Trudy GOIN 21/33.

—»— 1959: The formation of temperature conditions of the oceans (russ.). Gidrometeoizdat, Moscow.

SCHULEIKIN, V. V., 1932: Physics of the sea (russ.). Izd. AN SSSR.

SHEPPARD, P. A., 1958: Transfer across the earth's surface and through the air above. Quart. J. R. Met. Soc. 84, pp 205—224.

SIMOJOKI, H., 1946: On the temperature and salinity of the sea in the vicinity of the Bogskär lighthouse in the Northern Baltic. Soc. Sci. Fennica, Comm. Phys. — Math. XIII (7), pp 1—24.

—»— 1949: Niederschlag und Verdunstung auf dem Baltischen Meer. Fennia 71, No 1.

STURM, M., 1962: Über Methoden zur empirischen Berechnung der Hauptkomponenten des Wärmehaushaltes der Meeresoberfläche aus mittleren hydrometeorologischen Daten. Beitr. zur Meeresk., H. 9, pp 36—66.

SVERDRUP, H. U., 1937: On evaporation from the oceans. J. Mar. Res., vol. 1, No 1 pp 3—14.

—»— 1945: Oceanography for meteorologists. G. Allen and Unwin, London.

—»— 1946: The humidity gradient over the sea surface. J. Met., vol. 3, No 1, pp 1—8.

SWINBANK, W. C., 1959: Evaporation from the oceans. Sci. Rep. No 12, Contr. No AF 19 (604)—2179, The University of Chicago.

WITTING, R., 1908: Untersuchungen zur Kenntnis der Wasserbewegungen und der Wasserumsetzung in den Finnland umgebenden Meeren. I. Finnl. Hydr.-Biol. Komm. Unters. No 2.

—»— 1918: Hafsytan, geoidytan och landhöjningen utmed Baltiska hafvet och vid Nordsjön. Fennia 39, No 5.

WÜST, G., 1937: Temperatur und Dampfdruckgefälle in den untersten Metern über der Meeresoberfläche. Met. Zeitschr., Bd. 54, pp 4—9.

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